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VOLCANISM, PLUTONISM AND HYDROTHERMAL ALTERATION AT MEDICINE LAKE VOLCANO, CALIFORNIA

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ABSTRACT

Ion microprobe dating of zircons provides insight into the history of granitoid generation beneath Medicine Lake volcano (MLV), northern California. Part of the MLV geothermal system is hosted by a 319 ± 8 ka granodiorite intrusion that may be over 6 km in diameter. A sample of this intrusion was hydrothermally altered by 171 ka, indicating that the geothermal system was active by that time. Post-100-ka magmatism resulted in crustal melting, evidence for which is found in granitoid xenoliths erupted with numerous lava flows. Zircons within many of these xenoliths, analyzed by the ion microprobe (^{230}Th techniques) have ages less than 20,000 years, and as young as the age of eruption in the very latest Holocene. Sustained crustal input of basaltic magma generates the heat to melt crust, and acts as a continuing heat source to the shallow geothermal system.

INTRODUCTION

Medicine Lake volcano, located about 50 km ENE of Mt. Shasta in northern California, is a large Quaternary shield volcano that last erupted about 1000 years ago (Donnelly-Nolan et al., 1990). For nearly 30 years, it has been the site of geothermal exploration and represents one of the few U.S. prospects under consideration for new development (Hulen and Lutz, 1999).

Deep geothermal drillholes reveal that much of the exploitable geothermal system is hosted by hydrothermally altered granitoid rocks located ~ 2 km below the Medicine Lake caldera (Fig. 1). Granitoid xenoliths, mostly unaltered, are found within many Pleistocene and Holocene lava flows that cover the volcano (Mertzman and Williams, 1981; Grove et al., 1988; Lowenstern et al., 2000). In this paper we provide U-Pb and U-Th disequilibrium ages of zircon crystals ($ZrSiO_4$) extracted from granitoids obtained as drillcore and xenoliths. The resulting data provide a framework for understanding the intrusive history of the volcano and the genesis of its geothermal system. We provide evidence for the timing of intrusive episodes beneath the volcano and their relationship to ongoing volcanism.

BACKGROUND GEOLOGY

The current edifice of MLV has grown since about 0.5 Ma and overlies older lavas of the Modoc Plateau. Basement beneath the volcano is thought to be Sierran granite located at a depth of 10-20 km (Fuis et al., 1987). Eruptive products of the volcano cover about 2000 km^2 and are estimated at 600 km^3 in volume (Donnelly-Nolan, 1988), making MLV the largest volcano by volume in the Cascade Range.

MLV consists of a broad shield with a central 7 x 12 km caldera (Anderson, 1941). From the caldera region, the volcano's only ash-flow, the "andesite tuff" was erupted 171±43 ka (Herrero-Bervera et al., 1994). Argon ages of andesitic lavas forming the rim of the caldera are ~100-120 ka (Donnelly-Nolan et al., 1994). Since eruption of the ~85 ka Lake Basalt, nearly all volcanism within the Medicine Lake caldera has been silicic in composition (Donnelly-Nolan, 1988).

Around 11,000 B.P., voluminous basalts erupted to form the Giant Crater lava field, the basalt of Valentine Cave and other mafic lava flows scattered across the flanks of the volcano. This mafic episode was followed by a pause of about 6000 years without an eruption. Beginning about 5 ka, a variety of basaltic, intermediate and silicic lavas erupted across MLV, including several late Holocene silicic eruptions in and near the caldera. The Medicine Lake Glass Flow erupted within the caldera about 5000 B.P. Just to the west and east of the caldera, rhyolite eruptions took place respectively at Glass

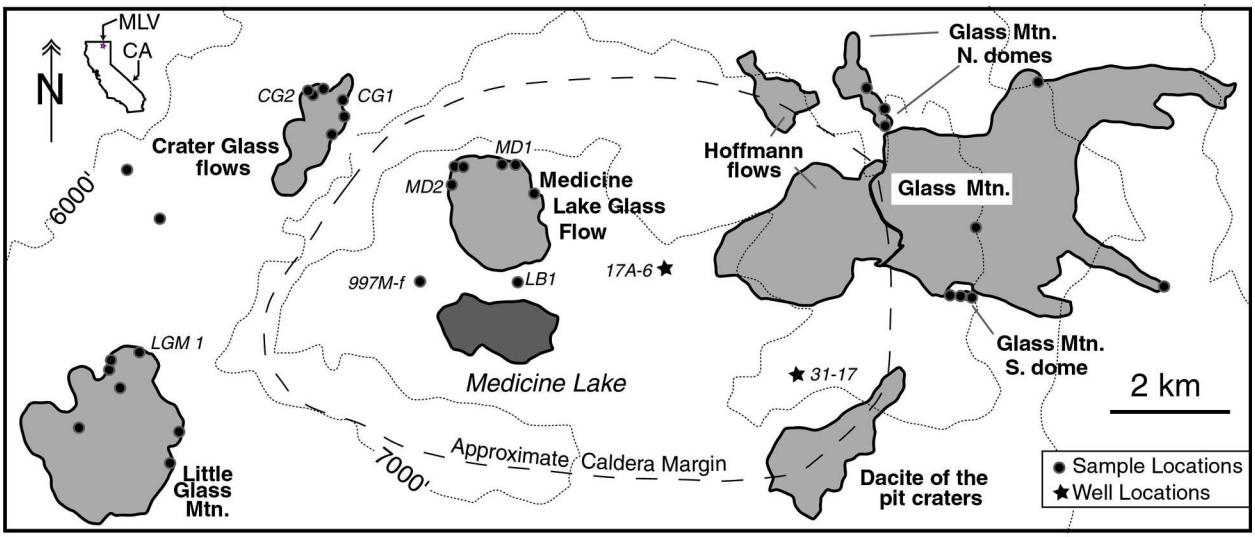

Fig. 1. Location map for samples from Medicine Lake volcano. Labeled samples are discussed in text. Unlabeled dots represent other studied xenoliths. Drill holes are shown by the stars.

Mountain (885 B.P.) and Little Glass Mountain (1065 B.P.) (Donnelly-Nolan et al., 1990).

In the 1980's, geothermal operators drilled four deep geothermal wells to depths ranging up to 2932 m. Three, including GMF 31-17, tapped a high-temperature, liquid-dominated geothermal reservoir and are capable of geothermal production (Hulen and Lutz, 1999). Another well, 17A-6, encountered commercial temperatures but was not flow tested because of mechanical difficulties (Hulen and Lutz, 1999).

SAMPLES

Granodiorite (64 wt.% SiO_2) from geothermal drillhole GMF 31-17 was made available by California Energy Corporation. The sample came from a depth of ~2570 m. This granodiorite body is located within the present-day geothermal system and shows clear evidence for hydrothermal alteration. In thin-section, mafic minerals have been converted to chlorite and Fe-oxides. Abundant, secondary, liquid-rich fluid inclusions give the feldspar phenocrysts a grainy appearance. Whole-rock analysis for oxygen isotope composition yielded a $\delta^{18}O$ of –1.5 ‰, revealing that abundant meteoric-derived water has interacted with the rocks, replacing the original magmatic oxygen.

Hydrothermally altered granitoid is also found continuously from 2348 m to 2933 m within geothermal drillhole 17A-6. Chemical analyses of drill cuttings of this material range in composition from 68 to 74 wt.% SiO_2 and thin sections display a range of micrographic to seriate to equigranular textures. Mafic minerals are altered to chlorite and secondary, liquid-rich fluid inclusions are abundant. Whole rock $\delta^{18}O$ on three samples revealed similar degrees of oxygen exchange as the sample from GMF 31-17 ($\delta^{18}O$ = –0.8 to –3.3 ‰).

A xenolith of hydrothermally altered granodiorite (68 wt.% SiO_2) was found within the 171 ka "andesite tuff". This unit is thought to have vented ~ 6km distant from the GMF 31-17 drilling site (Donnelly-Nolan and Nolan, 1986). The xenolith, sample 997M-f, is petrographically similar to the granitoids from the geothermal drillholes and has a whole-rock $\delta^{18}O$ of 0.5 ‰.

We also studied unaltered granitoid xenoliths found within younger volcanic units at MLV. The xenoliths range in composition from diorite to granite and include granophyric, seriate and equigranular textures. The fine- to medium-grain size, granophyric quench textures and vapor-rich fluid inclusions imply that these granitoids crystallized in the upper crust (< 8 km) where pressures are low and temperature gradients high. The xenoliths range from unmelted to highly melted (up to 50%). Presumably, the melting occurred during final transport to the surface within a hot magma column. Phenocrysts show no signs of hydrothermal alteration and $\delta^{18}O$ of whole-rocks and phenocryst separates display magmatic compositions from (+6 to +10‰). Sr isotopic compositions ($^{87}Sr/^{86}Sr$) of many of the xenoliths range up to 0.7050, far more radiogenic than basalts and basaltic andesites from MLV (Grove et al., 1988). That, and the elevated $\delta^{18}O$ imply a significant crustal component to the granitoid xenoliths (Grove et al., 1988; Lowenstern et al., in prep.) About 35 xenoliths have been collected and analyzed for their chemical

and isotopic compositions, but only six have been further prepared for geochronology (Table 1). They come mostly from the following Holocene lavas: Medicine Lake Glass Flow (MD1 and MD2), Crater Glass Flow (CG1 and CG2) and Little Glass Mountain (LGM1). In addition, we studied a single xenolith from the ~85 ka Lake Basalt (LB1).

Sample Name	Field Name	Wt.% SiO_2	$\delta^{18}O$ ‰	Host Eruption
GMF 31-17	GMF 31-17	63.9	-1.5	–
997M-f	997M-f	67.7	0.5	~171 ka
LB1	561M	71.0	9.3(q)	~85 ka
CG1	2050M	64.9	6.0	~ 1 ka
CG2	2049M	73.0	8.3	~1 ka
MD1	86-3	72.6	7.8	~ 5 ka
MD2	680M	74.2	8.8(q)	~ 5 ka
LGM1	2029M	73.7	8.8	~ 1 ka

Table 1. Granitoid samples from MLV. Oxygen isotopes for whole rocks, except where indicated (q = quartz separate). Host eruption = time of eruption that brought xenolith to surface.

SAMPLE PREPARATION AND ANALYTICAL TECHNIQUES

Zircons were separated with standard magnetic and heavy-liquid techniques and individual zircons were mounted in epoxy, ground nearly to half-thickness using 1500 grit wet-dry sandpaper and polished with 6- and 1-µm diamond abrasive. Each grain was photographed in reflected light and imaged with an SEM-based cathodoluminescence detector (Fig. 2). Isotopic data were acquired on the Stanford/USGS SHRIMP-RG (Sensitive, high-resolution ion microprobe – reverse geometry) at Stanford University. Grains older than 300,000 years were dated with the U-Pb isotopic system using techniques similar to Dalrymple et al. (1999). Ions were sputtered from zircons with a 10.5 nA primary O_2^- beam focused to a 25µm x 35µm spot. To minimize contamination by common Pb, the grain mount was cleaned in 1 N HCl acid for 5 minutes and the primary beam was rastered for 240 s on a 150-μm^2 region prior to data acquisition. The mass spectrometer was cycled eight times through peaks corresponding to Zr_2O, ^{204}Pb, background, ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{238}U, $^{232}Th^{16}O$ and $^{238}U^{16}O$.

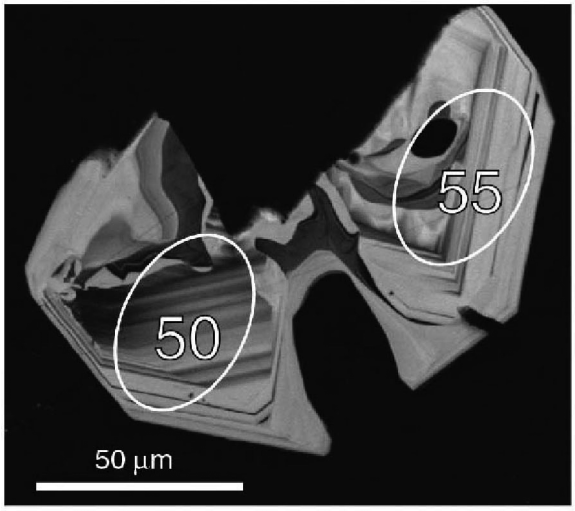

Fig. 2. Cathodoluminescence image of zircon crystal from LGM1. Dark bands correlate with areas high in U. The white circles represent areas analyzed with the SHRIMP-RG and yielded $^{230}Th/^{238}U$ ages of 50±10 and 55±10 ka (1 σ analytical errors). This crystal is among the oldest from this sample and may have been remobilized from an earlier intrusion.

For samples younger than 200,000 years, low radiogenic Pb and potential ^{230}Th–^{238}U disequilibrium makes U-Pb dating rather imprecise. Such young samples were thus dated by taking advantage of ^{230}Th–^{238}U disequilibrium systematics (Reid et al., 1997; Lowenstern et al., 2000). Because zircon incorporates U preferentially with respect to Th, a finite time (about 5 half-lives of ^{230}Th; i.e., ~375,000 years) must elapse before ^{230}Th is in isotopic equilibrium with its parent isotope, ^{238}U. Prior to that time, and especially for samples < 150,000 years old, the ^{230}Th–^{238}U system is effective as a geochronometer. Data were collected in ten scans per point for $^{90}Zr_2{}^{16}O$, $^{230}Th^{16}O$, background, $^{232}Th^{16}O$, ^{238}U, and $^{238}U^{16}O$ with count times per scan ranging from 2 to 15 s for each peak. Both the O_2^- beam and a 40-50 nA O^- beam were used to sputter the sample. Relative instrument sensitivity for ThO^+ and UO^+ was calculated by multiplying the measured $^{230}Th^{16}O/^{238}U^{16}O$ by a correction factor so that standard zircons from the 1.1 Ga Duluth Gabbro had a mean ($^{230}Th/^{238}U$) equal to unity. The correction factor was calculated during each run session and varied from 0.99 to 1.11. We calculated two-point model ages (Reid et al., 1997), representing the slope of a line connecting the zircon and whole-rock values on an activity diagram for (^{230}Th /^{232}Th) vs. (^{238}U /^{232}Th). The whole-rocks were analyzed by thermal ionization mass spectrometry with methods outlined

by Charlier (2000). Two-sigma uncertainties for individual zircon ages are generally ±10,000 years or greater, depending on the U concentration and U/Th.

RESULTS

U-Pb dating of zircons

Fourteen zircons from GMF31-17 (2570m) contained 300 to 2500 ppm U. On an inverse concordia diagram (Fig. 3), they plot along a line connecting a common Pb endmember with radiogenic Pb on concordia. Eleven grains from 997M–f, the hydrothermally altered xenolith, contained U concentrations from 1000 to 9500 ppm. Their higher U results in a greater proportion of radiogenic Pb, so that they plot on concordia (Fig. 3). The two zircon populations are colinear. A trend through the data, calculated with Isoplot/Ex (Ludwig, 2001) and fixed at the common Pb value for the whole-rock (~0.855), intersects concordia at 254±9 ka. This value must be corrected for initial U-Th disequilibrium because exclusion of Th from the zircon during crystallization results in less production of radiogenic Pb than in an equilibrium system (Wendt and Carl, 1985). We choose $(^{230}Th/^{238}U)_i = 0.33$, based on the average

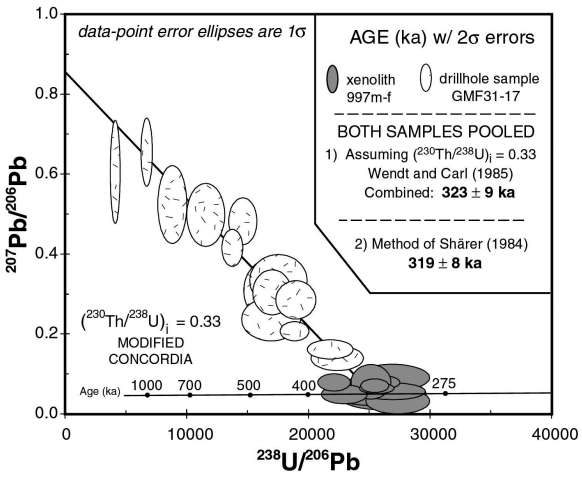

Figure 3. Modified Tera-Wasserburg diagram displaying U-Pb data for 25 zircons from two granitoids from Medicine Lake volcano. Error ellipses are 1 σ counting statistics. Because the data are colinear, they are treated as a single population. Calculated ages (with 2σ errors) in inset represent: 1) intercept with modified concordia (Wendt and Carl, 1985) assuming $(^{230}Th/^{238}U)_i = 0.33$ and 2) weighted mean of ^{207}Pb-corrected values for each spot, adjusted for initial $(^{230}Th/^{238}U)$ disequilibrium (Schärer, 1984).

U/Th of the zircons, the U/Th value for the whole-rock 0.362, and an assumption that the crystallizing melt was in isotopic equilibrium. The adjusted zircon crystallization age of the two (pooled) samples is 323 ± 9 ka.

The age of each zircon spot was also calculated individually, using the whole-rock common Pb value to perform a ^{207}Pb-corrected ^{206}Pb-^{238}U age. We then used the method of Schärer (1984) to estimate the extent of initial ^{230}Th disequilibrium for each spot. This correction adds from 50,000 to 85,000 years to the zircon ages, depending on the U/Th of the spot. The two samples yield statistically identical results (Fig. 4) and the weighted mean for data from both samples is 319±8 ka (2σ error: probability = 0.93; MSWD = 0.61).

U-Th disequilibrium dating of zircons

We analyzed 108 spots on zircon grains from six unaltered granitoid xenoliths. The calculated ages approximate the time of crystallization of individual zircons or zones within individual zircon grains. Zircons are refractory crystals, and U-Pb and U-Th disequilibrium ages are thus not re-set by magmatic temperatures or melting events (Zeck and Williams, 2002).

Fig. 4 lists weighted mean ages (and their 2σ standard deviations) for the zircons from each rock, excluding older grains that may represent earlier episodes of crystallization (of the same magma or a crystallized and remelted precursor). Four samples contained a preponderance of zircons with model ages less than 20,000 years (CG2, MD1, MD2, and LGM1). Besides the age of their zircon populations, these samples also share similar isotopic characteristics (whole-rock $^{87}Sr/^{86}Sr$, $^{143}Nd/^{144}Nd$, $^{208}Pb/^{204}Pb$ compositions) and exhibit pronounced whole-rock ^{230}Th-^{238}U disequilibrium (Lowenstern et al., in prep.). The four samples are colinear on a plot of $(^{230}Th/^{232}Th)$ vs. $(^{238}U/^{232}Th)$: the slope that connects them implies that these melts were generated 10,000 to 20,000 years ago. It appears that these four magmas were both generated *and* crystallized all within the past 25,000 years.

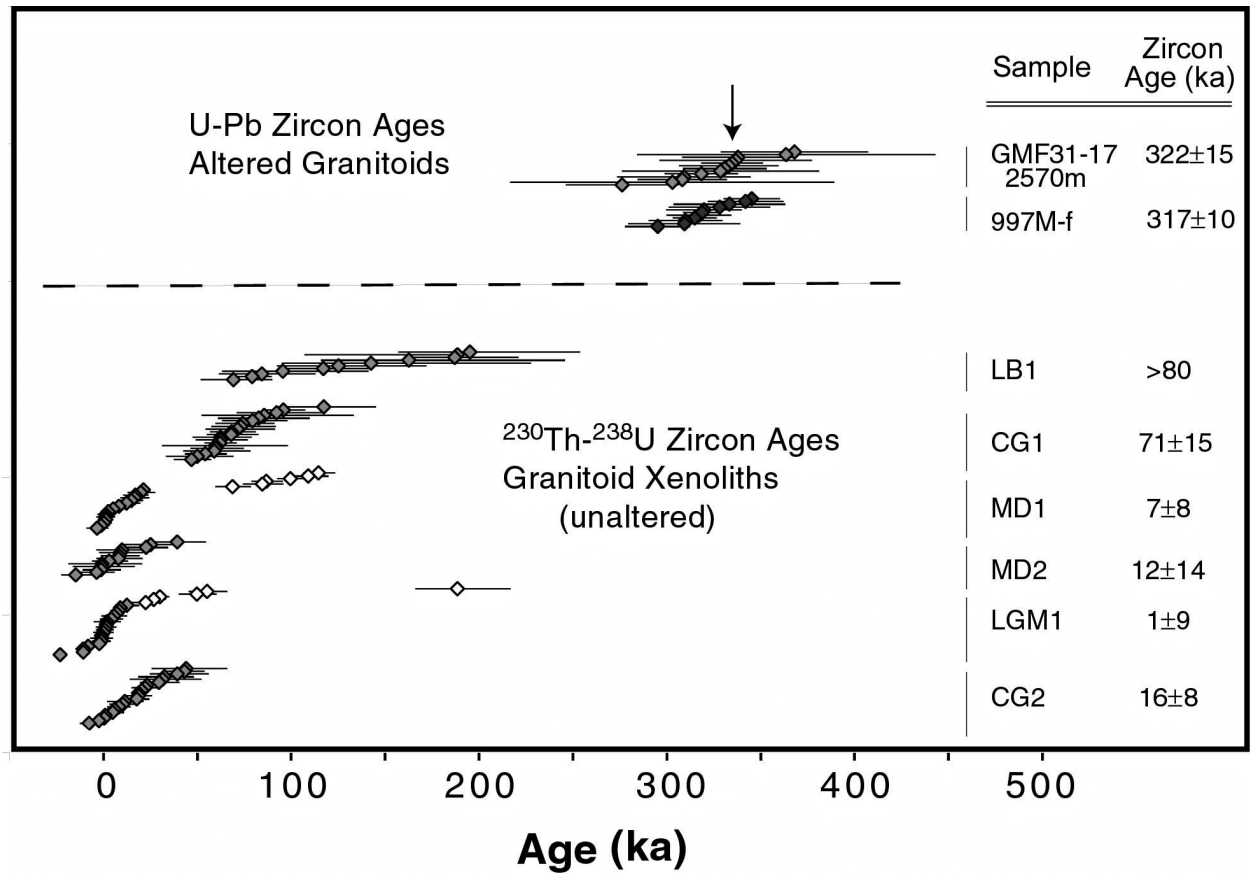

Fig. 4. Plot of ages for all zircon spots determined for this study. Each sample is shown separately, with oldest to youngest from top to bottom. The two samples at top were dated by U-Pb techniques, whereas data for the bottom six represent U-Th disequilibrium model ages (see text). Error bars are 1σ. Sample names and weighted-mean ages with their 2σ-standard deviations are shown at right. Only the zircons with filled symbols were included in the mean calculations, the others presumed to represent earlier crystallization events.

Two other xenoliths (CG1 and LB1) contain only zircons older than ~60,000 years. The former sample contains a relatively homogeneous zircon population with a weighted mean of 71,000±15,500 years, some 70,000 years prior to ejection of the xenolith during eruption of the Crater Glass flows. The latter granitoid contains zircons with a highly heterogeneous age spectrum; as young as its eruption age (~80-90 ka) and as old as 350 ka. Both xenoliths exhibit modest ^{230}Th-^{238}U disequilibrium, consistent with melt generation ~100 ka (Lowenstern et al., in prep.).

DISCUSSION

The identical U-Pb results for the hydrothermally altered granitoid from GMF31-17 and the xenolith 997M-f from the andesite tuff indicate that an intrusive episode took place at shallow depth beneath MLV at around 319 ka. Moreover, some 840 m of petrographically similar granitoid rocks in drillhole 17A-6 implies that this intrusion may be areally extensive (~ 6 km diameter). This would agree with earlier studies of gravity and seismic tomography data that an intrusion resides at shallow depths beneath a broad region of MLV (Finn and Williams, 1984; Evans and Zucca, 1988; Chiarabba et al., 1995).

By the time xenolith 997M-f was ejected during the 171 ka eruption of the "andesite tuff", it had become hydrothermally altered. Its δ^{18}O of 0.5 ‰ approaches the low negative values for altered rocks found in today's geothermal wells. It thus appears that the Medicine Lake geothermal system was active by ~170 ka. Plausibly, the emplacement of a large intrusion at ~320 ka provided the heat necessary to initiate the hydrothermal system.

It is noteworthy that most xenoliths are not hydrothermally altered, and similar unaltered granitoids have not been reported from the

exploratory drill holes. This strongly implies that unaltered intrusions reside beneath the bottom of the active meteoric-water-dominated geothermal system. Yet the vapor-rich fluid inclusions, fine-grained textures and petrographic characteristics imply a shallow origin for the xenoliths. Lowenstern et al. (2000) estimated that the unaltered xenoliths resided below the geothermal system and beneath the hydrothermally altered granitoid, at depths of 3-6 km, prior to stoping, ascent and eruption within young magma. Because the xenoliths lack annealed fractures containing secondary liquid-rich inclusions, and show no evidence for alteration, these rocks likely were never subjected to temperatures below the ductile-brittle transition (~400°C). As such, fluids from the superjacent geothermal system may not have had access to the ductile and impermeable region where these granitoids resided. This implies that below the geothermal system, the thermal gradient may increase rapidly and locally reach near-magmatic temperatures at mid-crustal depths.

Four unaltered xenoliths contain zircons that crystallized less than 20 ka. They were erupted at three different locations over a relatively wide (~10 km) geographic area. Apparently, a variety of small-volume magmas were generated over the last 20,000 years beneath Medicine Lake volcano and rapidly crystallized prior to their mobilization and eruption within juvenile magma. The xenoliths are isotopically distinct from the erupted magmas that brought them to the surface. Lowenstern et al. (2000), using some of the present data set, argued that the xenoliths were generated by localized crustal melting during periods of high-flux throughput of mafic magma. Their distinct major- and trace-element compositions and geographic distribution argues for localized melting over a broad region. It is possible that this melting occurred during the large mafic event at ~11 ka that resulted in eruption of the Giant Crater and basalt of Valentine Cave flows.

An alternate mode for viewing the data is a single histogram for all 133 zircons analyzed, which is useful for identifying discrete age populations. As expected, there is an obvious grouping at 318 ka, associated with the intrusion found in GMF 31-17 and the altered xenolith (997M-f). Zircons from the younger xenoliths display three possible discrete episodes of crystallization, the most recent one indistinguishable from the late Holocene time of eruption. The peaks at 19 and 61 ka may represent intrusive/crystallization episodes not specifically correlated with extrusions (Lowenstern et al., 2000). Alternatively, many of the youngest zircons (3 and 19 ka peaks) could have formed after melting associated with the input of mafic magma at ~11,000 B.P., but are difficult to identify given the relatively large errors for individual zircon analyses.

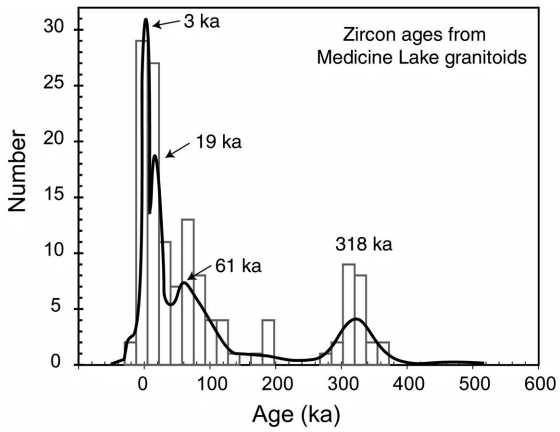

Fig. 5. Histogram for zircons from all units. Overlain is a calculated age-density curve with peaks chosen so that the MSWD around each mean approaches one (calculated by Isoplot/EX; Ludwig et al., 2001). Young peaks are present at 3 ka, 19 ka and 61 ka. The older peak at 318 ka represents zircons from the hydrothermally-altered drill core (GMF31-17) and xenolith (997M-f).

SUMMARY

Over the past few hundred thousand years, a variety of granitoid magmas have intruded under Medicine Lake volcano. We find evidence for a potentially large intrusion (> 6 km in diameter) emplaced at ~ 320 ka. This intrusion hosts much of the present geothermal system and is hydrothermally altered. The alteration, and thus the geothermal system, appears to have been generated prior to ~170 ka.

Ongoing volcanism of mafic-to-intermediate composition lavas at MLV over the past 100,000 years resulted in partial melting of the crust, causing generation of small granitoid bodies. These intrusions were ultimately emplaced at 3-6 km depths, below the intrusion that hosts the hydrothermal system. Four out of six of the studied xenoliths apparently were generated and crystallized within the past 25,000 years. The xenoliths are not chemically equivalent to the host lavas that brought them to the surface, though they may represent an important assimilant in the generation of MLV rhyolites (Grove et al., 1988). The process of crustal melting and subsurface crystallization is an important aspect of heat transfer from mafic magma to crustal geothermal systems. At locations where basalt is rapidly transferred from mantle to surface, crustal

hydrologic systems may receive little heat input. At MLV, heat is transferred from the mafic magma system to generate silicic magmas, some of which crystallize and lose heat at shallow crustal depths. This process ultimately fuels the geothermal resource found at volcano-hosted hydrothermal systems (Duffield et al., 1994).

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REFERENCES

Anderson, C.A. (1941) "Volcanoes of the Medicine Lake Highland, California" *University of California Publication, Bulletin of the Department of Geological Sciences*, **25**, 347-422.

Charlier, B.L.A. (2000) "U-Th isotopic constraints on the pre-eruptive dynamics of large-scale silicic volcanism: examples from New Zealand", *Ph.D. Thesis, The Open University*, 393p.

Chiarrabba, C. and Amato, A. (1995) "Variations on the NeHT high-resolution tomography method: A test of technique and results for Medicine Lake volcano, northern California" *Journal of Geophysical Research* **100** (B3) 4035-4052.

Dalrymple, G.B. *et al.* (1999) "Age and thermal history of The Geysers plutonic complex (felsite unit), Geysers geothermal field, California: A $^{40}Ar/^{39}Ar$ and U-Pb study" *Earth and Planetary Science Letters* **173**, 285-298.

Donnelly-Nolan, J.M. (1988) "A magmatic model of Medicine Lake volcano, California" *Journal of Geophysical Research*, **93**, 4412-4420.

Donnelly-Nolan, J.M. and Nolan, K.M. (1986) "Catastrophic flooding and eruption of ash-flow tuff at Medicine Lake volcano, California" *Geology*, **14**, 875-878.

Donnelly-Nolan, J.M. *et al.* (1990) "Post-11,000-year volcanism at Medicine Lake volcano, Cascade Range, northern California" *Journal of Geophysical Research*, **95**, 19693-19704.

Donnelly-Nolan, J.M. *et al.* (1994) "Incomplete extraction of radiogenic argon from high-silica andesites: Implications for K-Ar dating" *ICOG Abstract Volume, USGS Circular* **1107**, 84.

Duffield, W.A., Sass, J.H., and Sorey, M.L. (1994) "Tapping the Earth's Natural Heat" U.S. Geological Circular 1125, 63 p.

Evans, J.R. and Zucca, J.J. (1988) "Active high-resolution seismic tomography of compressional wave velocity and attenuation structure at Medicine Lake volcano, northern California Cascade Range" *Journal of Geophysical Research* **93** (B12) 15,016-15,036.

Finn, C. and Williams, D.L. (1984) "Gravity evidence for a shallow intrusion under Medicine Lake volcano, California" *Geology* **10**, 503-507.

Fuis, G.S., (1987) "A geologic interpretation of seismic-refraction results in northeastern California" *Geological Society of America Bulletin* **98**, 53-65.

Grove, T.L. *et al.* (1988) " Assimilation of granite by basaltic magma at Burnt Lava flow, Medicine Lake volcano, northern California: Decoupling of heat and mass transfer" *Contributions to Mineralogy and Petrology* **99**, 320-343.

Herrero-Bervera, E. et al. (1994) "Age and correlation of a paleomagnetic episode in the western United States by 40Ar/39Ar dating and tephrochronology: The Jamaica, Blake or a new polarity episode?" *Journal of Geophysical Research* **99**, 24091-24103.

Hulen, J.B., and Lutz, S.J. (1999) "Altered volcanic rocks as hydrologic seals on the geothermal system of Medicine Lake volcano, California" *Geothermal Resources Council Bulletin* **28**, 217-222.

Lowenstern, J.B., *et al.* (2000) "U-Th dating of single zircons from young granitoid xenoliths: new tools for understanding volcanic processes" *Earth and Planetary Science Letters* **183**, 291-302.

Ludwig, K.R. (2001) "Users Manual for Isoplot/Ex version 2.49" *Berkeley Geochronology Center Special Publication* **1a**, 56 p.

Mertzman, S.A., and Williams, R.J. (1981) "Genesis of Recent silicic magmatism in the Medicine Lake Highland, California: evidence from cognate inclusions found at Little Glass Mountain" *Geochimica et Cosmochimica Acta* **45**, 1463-1478.

Reid, M.R. *et al.* (1997) Prolonged residence times for the youngest rhyolites associated with Long

Valley Caldera: "^{230}Th -^{238}U ion microprobe dating of young zircons", *Earth and Planetary Science Letters*, **150**, 27-39.

Schärer, U. (1984) "The effect of initial ^{230}Th disequilibrium on young U-Pb ages: the Makalu case, Himalaya" *Earth and Planetary Science Letters* **67**, 191-204.

Wendt, I., and Carl C. (1985) "U/Pb dating of discordant 0.1 Ma old secondary U minerals" *Earth and Planetary Science Letters* **73**, 278-284.

Zeck, H.P., and Williams, I.S. (2002) "Inherited and magmatic zircon from Neogene Hoyazo cordierite dacite, SE Spain – Anatectic source rock provenance and magmatic evolution" *Journal of Petrology* **43**, 1089-1104.